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Palaeomagnetic investigations of sediments cores from Axios zone (N. Greece): implications of low inclinations in the Aegean

E. Aidona[1], **D. Kondopoulou**[1], **R. Scholger**[2], **A. Georgakopoulos**[3], **and A. Vafeidis**[4]

[1]Dept. of Geophysics, School of Geology Univ. of Thessaloniki, Greece
[2]Paleomagnetic Laboratory Institute of Geophysics – University of Leoben, Austria
[3]Dept. of Mineralogy-Petrology-Economic Geology, School of Geology, Univ. of Thessaloniki, Greece
[4]Department of Mineral Resources Engineering, Technical, University of Crete, Chania, Greece

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Correspondence to: E. Aidona (aidona@geo.auth.gr)

Abstract

Sediment cores from 13 deep boreholes (1–4.1 km) from Axios zone in Northern Greece have been studied by means of palaeomagnetism. Both low field magnetic susceptibility and intensity of the natural remanent magnetization (NRM) indicate rather weakly magnetised materials. 390 samples have been subjected to demagnetization process (AF and thermal) revealing in most of the cases the presence of magnetite. Isothermal remanent magnetization (IRM) acquisition curves and thermomagnetic analysis suggest the dominance of magnetite. 30 thin sections were studied in order to more precisely characterise the magnetic mineralogy of the samples. This investigation also reveals the presence of magnetite and pyrite in framboidal form. An attempt to re-orient some of the samples was partially successful by using the viscous component and the anisotropy method. Re-orientation techniques were applied in order to correct the palaeomagnetic directions due to the orientation ambiguity of the core samples. The palaeomagnetic results confirm the clockwise Cenozoic rotation, in the study area in agreement with the overall pattern of the onshore results from previous investigations.

Finally, the observed inclinations of characteristic remanences in these rocks are much lower than the expected ones but converge with those obtained from formations on land.

1 Introduction

Numerous palaeomagnetic studies in Greece have contributed to the better understanding of the geotectonic framework of the area and to the establishment of magnetostratigraphic columns (Duermeijer et al., 2000; Kondopoulou 2000, and references therein; Van Hinsbergen et al., 2005). In particular, the presence of stable magnetization components of different ages is indicated from previous palaeomagnetic and magnetostratigraphic studies in various formations (ophiolites, plutonites, sediments) on and around the Chalkidiki peninsula in Northern Greece (Fig. 1) (Kondopoulou and Westphal, 1986; Edel et al., 1991; Feinberg et al., 1994; Kondopoulou, 1994; Haubold et al., 1997). These components imply the presence of clockwise rotation for the Cenozoic, consistent with the general pattern of the broader area. On the other hand, inclination values are usually lower than expected by 5°–20° (Kondopoulou and Westphal, 1986; Kondopoulou et al., 1996; Beck et al., 2001). However, it should be noted that these low inclinations have been observed only at surface outcrops or borehole samples from shallow coring of lake sediments (Papamarinopoulos, 1978). Preliminary results from early palaeomagnetic and rock magnetic investigations from four deep-drill cores (Kass1, Kass2, Kass3, Kass4) have already been published in a previous paper (Aidona et al., 2001).

In the present study we present palaeomagnetic information from a large number of core samples from 13 deep boreholes (up to 4000 m) that were drilled down by the Greek Petroleum Company for oil exploration. It must be clarified that in these boreholes core sampling is sparse and is limited only on few specific strata of interest. The vast drilling procedure produces only small fragments ("cuttings"). Nevertheless, the dataset which was obtained provides valuable information both on the validity of the rotational pattern and the long-standing problem of the low inclination values in the broader area.

2 Geological setting and sampling

The study area includes the north (Giannitsa-Thessaloniki plain) and eastern (West Chalkidiki and Kassandra peninsula) margins of Thermaikos gulf. From the geotectonic point of view the area belongs to the Axios zone (internal Hellinides). Mesozoic rocks (limestones, ophiolites) of Axios zone form the basement of the area. During Paleogene an extensive NNW-SSE trending trench was formed and filled by Paleogene molasic sediments (sandstones, few limestones), while in Miocene a new NNW-SSE graben was formed again and filled gradually during Neogene-Quaternary by mainly clastic sediments (sands, sandstones, clays, red beds) and few marls and limestones (Antonios Fm, Triglia Fm, Trilofos Fm, Moudania Fm) (Syrides, 1990). At present, basement and Paleogene sediments expose at the mountainous periphery of the Thermaikos gulf while in the low relief hilly area the basement is covered by the Neogene – Quaternary deposits exceeding 4000 m in thickness.

Axios area (drills Kor-2, Al-1, Lo-1) belongs to the Paiko zone. This zone consists of volcano-sedimentary rocks of Upper Jurassic age, a calcareous horizon of the same age and flysch of Low Cretaceous age which is covered by Middle-Upper Cretaceous sediments (Mercier, 1968).

In order to simplify descriptions the sampled cores are grouped into 4 sets according to their position: Axios area (Kor-2, Al-1, Lo-1), Kassandra peninsula (Kass1-2-3-4, Po-1) Epanomi (Ep-1, Ep-2, Ep-b1, Ira-2) and a submarine core, Nireas-1 (Fig. 1).

The stratigraphy of the drilled cores is rather simple: all 4 groups contain a thick clastic sediment layer, which covers the pre-Neogene basement (Fig. 2). In particular, in the Kassandra set, the clastic layer (Eocene-Miocene) overlays the ophiolite basement. In Epanomi set the clastic layer of Eocene-Pliocene age covers the Jurassic limestone and in Axios set the same clastic sediments (Eocene-Pliocene) overlay the Mesozoic metamorphic basement (Georgala, 1990, 1994).

From all these deep boreholes, 620 core samples from various depths were obtained. In the laboratory 390 of these samples were demagnetized using thermal and alternating field demagnetization. Tilt values have been measured only for Kass3, and Kass4 boreholes and suggest limited tilting (<15°) of the samples units, with only few exceptions reaching tilts up to 25°.

3 Laboratory treatment

3.1 Palaeomagnetic measurements

The magnetic susceptibility from all 620 specimens was measured with a Bartington MS2 device at low frequency (0.47 kHz). The mean susceptibilities as well as the mean intensities values for the 13 boreholes are displayed in Table 1. The mean susceptibility for the samples of Epanomi group shows negative values, which indicate the presence of paramagnetic minerals into their structure. The majority of the other boreholes show a mean susceptibility, ranging from 12–18 X SI^{-5} units and only Kor-2 gives mean susceptibility values up to 28 X SI^{-5} units.

The natural remanent magnetization (NRM) of all 620 specimens was initially measured on a Molspin spinner magnetometer. The samples from Epanomi group turned to be very weakly magnetized (around 0.2 mA/m) thus they were measured on a cryogenic magnetometer. All the remaining samples showed higher intensities of magnetization, up to 20 mA/m (Kor-2) (Table 1).

Magnetic cleaning was performed in 390 specimens by using thermal stepwise (Th) with 50°C step and alternating field (AF) with 5 mT–10 mT step, demagnetizations.

3.2 Rock magnetism and mineralogical analysis

Isothermal remanent magnetization (IRM) and thermomagnetic analysis have been performed on 70 specimens for the determination of the magnetic carriers. The IRM was imparted using a pulse magnetizer and was measured on a spinner magnetometer. The majority of the thermomagnetic analyses were performed on a KLY-3 Kappabridge which measures the variation of the magnetic susceptibility with temperature. Only for few samples the Susceptibility Temperature Bartington Device was used.

30 polished thin sections were prepared and were studied by reflected light microscopy under high magnification, in order to investigate the distribution of the minerals into the samples. The morphology of the minerals was studied by a GEOL-840A

Scanning Electron Microscope, equipped with a LINK ISIS microanalyzer and using carbon coated samples.

4 Results

The majority of the samples of Kassandra and Axios areas were weakly magnetized but stable components have been isolated in most cases. The thermal stepwise demagnetization has been successful up to 400°–450°C. Above this temperature, the intensity of the magnetization consistently increased in the majority of the samples (Fig. 3a). This behavior is probably related to alteration (oxidation) of pyrite, which is present in all samples and to the formation of new parasitic minerals which overlap the high coercivity component of magnetization of the samples. Thus, the high temperature component has been lost and only a stable low temperature component (between 200°–450°C) has been isolated. In reverse, the AF demagnetization process proved to be more successful for the determination of the magnetic components, as during the AF procedure generation of new parasitic minerals is avoided (Fig. 3b). Therefore, the characteristic component of magnetization was possible to isolate in most of the samples in fields between 20–80 mT. The thermal and AF demagnetizations were performed in twin specimens, so the high coercivity component calculated by AF was used as a reference for checking the stable direction obtained through the thermal procedure. The criteria used to decide whether or not a direction would be considered as reliable were: a) a smooth decay curve, b) no variation of susceptibility during heating, c) alignment of at least 4 points, and e) a reliable a_{95} value (<15°). The characteristic directions of the magnetization were determined by least squares fitting (Principal Component Analysis) through selected points. All samples from Epanomi area were rejected due to their extremely low intensity of magnetization.

IRM measurements reveal the dominance of magnetite as the main magnetic carrier in most of our samples (Fig. 4). Many samples achieve their saturation remanence in fields of 200–300 mT. In few cases the saturation is not reached up to 1200 mT, thus indicating the presence of hematite. Thermomagnetic analysis confirms the dominance of magnetite, as the maximum blocking temperature reaches 580°C in most of the cases. It is also clear that during the heating, secondary magnetite is formed (Fig. 5).

Finally, the study of the thin sections confirms the presence of magnetite, hematite and pyrite and reveals the presence of ilmenite and chromite (Fig. 6). Magnetite is considered as a detrital mineral coming from the surrounding igneous and metamorphic rocks deposited in the sedimentary basin. Pyrite microcrysts appear mainly in the framboidal form but clouds of pyrite microcrysts not organized in framboids are also present (Fig. 6). The framboids are usually found in colonies, filling, in some cases, fossils of micro-organisms (Fig. 6). The presence of framboidal pyrite is attributed to low temperature and neutral to alkaline pH values. These conditions reduce pyrite solubility and are essential for the formation and following preservation of the framboids (Rickard, 1969, 1970).

5 Re-orientation of cores

In order to derive palaeomagnetic results from cores it is necessary, as a first step, to bring back core pieces into their initial position with respect to north and to vertical. For this reason, in the last decades, a number of new techniques of re-orientation of cores have been developed, and, among others, the palaeomagnetic reorientation. This method is based on the isolation of the viscous component of the magnetization which records the direction of the present geomagnetic field (Fuller, 1969; Van der Voo and Watts, 1978; Shibuya et al., 1991; Hailwood and Ding, 1995). According to this method we isolate the viscous component from our samples but as it is shown in a stereographic projection in Fig. 7 the majority of the samples show very shallow inclination values. This fact leads us to the conclusion that it is impossible to use these values in order to re-orient our cores. It seems that a parasitic magnetization which was created into the stockroom (Athens storage center) had affected the samples. Thus, storage in horizontal position of the samples for several years leads to the destruction of their viscous component of the magnetization. Samples from Axios cores seem to be the exception (Fig. 8). As these samples were stored in another stockroom (Kavala storage center), they have obviously not been affected by the same parasitic magnetization.

Regarding the position of the cores with respect to the vertical, the method of anisotropy of the magnetic susceptibility was applied (Hailwood and Ding, 1995). By measuring the anisotropy of magnetic susceptibility we could determine the plane of k_{max} and k_{int} and the position of the k_{min} axe of anisotropy. For a sedimentary sequence which was deposited in an undisturbed environment, the k_{min} axe would be perpendicular to the bedding plane (Hamilton and Rees, 1970; Tarling and Hrouda, 1993). Measurements of anisotropy of the magnetic susceptibility were performed in 120 specimens and the results are shown in Fig. 9. Only specimens with values of magnetic foliation higher than 1.010 were considered reliable and among these, samples with k_{min} inclination values (blue circles) less than 75° were corrected. Thus, a small set of samples (10% of the total) with corrected declination and inclination values was derived (Fig. 10). A large dispersion is observed for both declination and inclination values but within reasonable limits of reliability. Nevertheless, some general trends can be safely recognized and will be discussed in the next paragraph.

6 Discussion

In spite of the generally weak magnetizations the careful palaeomagnetic and rock magnetic studies of the cores were successful in isolating stable components in most cases. The obtained magnetic parameters can be exploited mostly as far as the inclination values are concerned. In Table 2 all the obtained results from this study are presented. The criteria used for the acceptance or rejection of the samples, are based on inclination values, a_{95} values and the number of specimens measured from every core sample. It is important to notice that the final grouping of the results was based on their age distribution.

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A mean inclination was computed using the method of Enkin and Watson (1996), which disregards all declination data. Inclination values have been compiled for the Eocene-Oligocene and separately for the basement. Results are as follows: $I=32.9^{+5.6}_{-5.6}$, $\kappa=10$, for the Oligocene, $I=44.3^{+5.2}_{-4.7}$, $\kappa=9$ for the Eocene and $I=31.1^{+7.9}_{-6.9}$, $\kappa=10$ for the ophiolite basement (Jurassic?).

Inclination data from an allochthonous ophiolite cannot be interpreted unless the palaeohorizontal can be determined. This is practically impossible for drill core materials. In order to overcome this difficulty and properly evaluate the observed inclination values, we have compiled all published data from onshore formations from the broader area (Table 3a). The general agreement of inclination values from the two datasets is supported by one particular case. Palaeomagnetic results from ophiolites and associated sediments in the edge of Kassandra peninsula yield a direction with D=325.4° and I=34.1° for the Mesozoic component (Feinberg et al., 1996). Given that in this case the palaeohorizontal was unambiguously determined, it is clear that this inclination value (34.1°) can be used as a reference value for inclination data from the neighboring area.

In Fig. 11 the distribution of inclination values for Kassandra area is shown. It is observed that Eocene inclination values for drill cores Kass4 and Pos are much higher than for the other boreholes. This difference indicates the presence of local intense tectonic events (fast block tiltings) during the end of Eocene. This fact is in good agreement with the formation of the sedimentary basin and its rapid extension after Middle Eocene as confirmed by the sedimentological study (Roussos, 1994). If we extend the broader area to a largest one, up to the western edge of the Pelagonian zone a small dataset with published directions (Table 3b) displays a range of inclinations comparable with the low inclinations calculated in the present study but divergent from the observed high inclinations in the two drill cores Kass4 and Pos. Also, the eastern part of the study area (Sithonia and central Chalkidiki) gives similar low inclination results for Eo-Oligocene plutonic formations (I=31°). Therefore, it can be derived that the overall inclination is low and the high inclinations in Kass4 and Pos are strictly limited to the internal part of the Thermaikos gulf possibly due to local tectonism related to the formation of the graben of Thermaikos gulf.

The expected inclination values computed from the European apparent polar wander path are 51° (Westphal et al., 1986) or 53.7° (Besse and Courtillot, 2002) for the Eocene and 56° (Westphal et al., 1986) or 54° (Besse and Courtillot, 2002) for the Oligocene. Taking these values into account the observed inclination values are consistently low (31°/44°) for Eocene-Oligocene formations or Eocene overprints (Table 3a). The comparison of the two datasets clearly shows that inclination values from both onshore and borehole formations are compatible and much lower than the expected values (by ~20°). A second dataset, comprising results from reoriented samples (Fig. 10) should be examined with due care. In this small group (N=25) the mean declination value is in a good agreement with almost all published on-shore results for the broader Chalkidiki area (Table 3a). As far as the inclination values are concerned their mean (I=58.2°) is close to the expected ones for the area and the Eocene-Oligocene period. At a first approach this could be in contradiction with the generally lower inclination values measured in the present dataset but also in the published in-shore data. This divergence is not representative since the data used for the reorientation are derived from the cores Kass4, Pos and Kor-1 where high inclination values prevail. Therefore, the mean inclination value for reoriented samples is shifted to a higher range. For all these reasons, we consider as representative the inclinations obtained from the totality of the studied samples.

The problem of low inclinations for the Cenozoic formations in the broader Aegean area has been widely discussed (Kissel and Laj, 1988; Van der Voo, 1993; Beck and Schermer, 1994; Beck et al., 2001; Kissel et al., 2003). The thorough examination of this problem by Beck et al. (2001) led to the suggestion of a NW motion of the Aegean block by ~500 Km with respect to N. Europe, which cannot be supported by geological data. In their discussion on shallow inclinations in the Aegean realm, Kissel et al. (2003) conclude that the problem is not fully understood. A different approach used by Krijgsman and Tauxe, (2004) based on the "elongation/inclination" method of Tauxe and Kent (2004), led the authors to the suggestion that an inclination error is the main responsible for low inclinations in sedimentary formations and inappropriate tilt corrections for the ones observed in lava flows in the area. This last assumption is based on the Beck et al. (2001) study of volcanic rocks in Lesvos island. As far as the sediments are concerned their conclusion seems realistic and converges with the one of Kissel et al. (2003). On the other hand, we believe that "inappropriate tilt" is a very simplistic approach for lavas, as similar low inclinations in volcanics, have been reported from other Oligo-Miocene formations in the broader Aegean (Kissel et al., 1986a, b; Morris, 1995, 2000; Haubold et al., 1997). It is difficult to consider that in such a broad dataset obtained by different groups and with various sampling techniques, a systematic error in applying tilt corrections has occurred.

Several studies have revealed the same anomalous inclinations further to the region of central Asia (Hankard et al., 2007, and references therein). Chauvin et al. (1996) proposed that inclination anomaly progressively increases from 0° on the Atlantic margin to about 10° in the eastern Mediterranean and Middle East to reach maximum of 25° in central Asia, due to a regional non-dipole field. Van der Voo and Torsvik (2001) suggested that an octupole field may have been responsible for the palaeolatitude anomalies in central Asia, adopting the earlier suggestion by Westphal (1993). Finally Si and Van der Voo (2001) proposed that the large discrepancies between observed and predicted palaeolatitudes in Asia during latest Cretaceous and Tertiary can be explained by the contribution of a long-term non-dipole field to the total time-averaged geomagnetic field.

In the present study, as the on-shore data come mostly from plutonic rocks and are in good agreement with data from the borehole samples (both have shallow inclinations), we consider that inclination flattening does not contribute significantly to the observed inclination anomalies. Additionally, if the Eocene alternative pole for Eurasia suggested by Westphal (1993) is taken into account (I_{exp}~35°) a satisfactory match is observed with the inclination values reported here, suggesting that large-scale crustal displacements are unnecessary for this geological period.

7 Conclusions

The study of drill sediments cores from Axios zone provided a set of meaningful results for an area and a time span in which such data are very scarce (Eocene – Oligocene).

In the case of Epanomi samples, possible dolomitization could have affected these sediments, leading to their remagnetization. As a result the total magnetic grains were replaced by paramagnetic ones destroying the remanent magnetization of the samples.

The usage of the viscous component of the magnetization for the correction of declination proved to be useless due to storage effect. Only in few cases the method was applied successfully and reorientation of declination (for these specific samples) was possible. The reorientation procedure might have proved more successful if data from the totality of the boreholes could be used.

The calculated mean inclination values seem to be in good agreement with the overall pattern of inclinations of the broader area from the on-shore formations. The total set of inclination values is lower than the expected ones following the magnetic pole of Eurasia.

The observed low inclinations in plutonic rocks and sediments in previously studied on-shore formations, in combination with the results from the present study lead us to conclude that it is difficult to accept the sediment compaction as the dominant mechanism for low inclination. For the particular time range (Eocene-Oligocene) the alternative pole of Westphal (1993) could possibly give a satisfactory interpretation.

Nevertheless, the overall problem of shallow inclinations in the broader area remains still an open issue. Additional data from well-dated volcanic rocks of similar ages with severe control on stratigraphic corrections would be welcomed in order to elucidate the problem.

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Table 1. Mean values of intensity and magnetic susceptibility for all studied cores.

Name	N	Mean Intensity Value (mA/m)	Mean Susceptibility Value (SI 10^{-5})
EP-1 (Cores1-2)	10	1.138	18.17
EP-1 (Cores 3–7)	24	0.137	−0.366
EP-2	21	0.081	1.127
EP-B1	36	0.196	−0.113
IRA-2	65	0.012	−0.377
KASS-1	56	2.43	16.1
KASS-2	66	0.88	11.93
KASS-3	59	1.34	14.02
KASS-4	91	1.51	16.26
PO-1	119	5.32	17.92
Kor-2	26	26.43	28.08
Al-1	14	8.41	15.84
Lo-1	8	0.34	2.97
Nir-1	26	2.68	14.43

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Table 2. Obtained results from the present study. Name = name of the borehole, I = inclination value, a_{95}=95% cone of confidence, k = Fisher's precision parameter, n = number of specimens, A = accepted, R = rejected (see in the text for details).

Name	I	a_{95}	k	n	Comment
Kass-1					
1117–1120	10.8	4.3	73.39	12	A
1160	23.1	40.9	165.3	3	R
1190	3.6	42.3	155.4	3	R
1443	27.3	40.4	8.87	3	A
1544–1547	31.8	11.5	21.8	7	A
1758	24.6	12.6	52.75	4	A
Kass-2					
1133–1140	30.8	10.9	10.83	13	A
1198–2005	23.8	9.7	16.58	11	A
1464	39.3	–	5.53	3	R
Kass-3					
1688–1694	33.7	9.2	11.68	16	A
1867–1869	38.5	39.8	3.83	5	A
1971–1974	35.3	17.3	8.3	8	A
Kass-4					
1863–1869	21.8	11.1	12.69	11	A
2328–2335	23.6	16.5	7.71	9	A
2650–2655	77.1	4.0	1348	5	A
2796–2804	46.4	12.8	18.07	7	A
2902–2908	71.9	9.4	59.47	5	A
3040–3043	20.0	14	18.99	6	A
Pos-1					
1225–1231	28.1	11.4	10.94	12	A
2642–2648	24.8	12.1	8.17	14	A
3027–3032	26.9	23.9	5.5	7	A
3287–3290	64.0	13.5	21.57	6	A
3467–3472	62.5	8.0	29.5	10	A
4116–4118	38.3	19	14.66	5	A
Kor-2					
2704–2705	74.5	24.9	34.16	4	A
2742–2787	73.3	8.1	82.07	5	A
2796–2798	48	12.9	10.19	11	A
Nir-1					
2041–2047	30.2	25.7	13.19	4	A
2785–2787	19	12.6	53.34	4	A
3563–3567	65	5.3	48.33	13	A
Al-1					
1623–1626	27.1	18.3	11.37	6	A
1700–1704	53.2	19.6	7.27	8	A
Lo-1					
3011–3015	48.4	46.8	3.37	5	A

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Table 3. (a) Published data for the broader area.

Area	Formation	Age	N	D	I	A_{95}	Reference
1. Chalkidiki (Kassandra)	Sediments	Mio-Pliocene	9	24.7	54.2	5	Haubold et al. (1999)
2. Axios	Sediments	Miocene	4	20	46	17	Aidona et al. (1996); Sen et al. (2000)
3. Strimonikos	Plutonics	Oligocene-Miocene	2	026	47	–	Westphal et al. (1991)
4. Chalkidiki (Paliouri)	Sediments	Eocene – Oligocene	4	220.2	−47.3	–	Feinberg et al. (1994)
5. Chalkidiki (Sithonia-Ouranoupoli)	Plutonics	Eocene – Oligocene	8	037	31	9	Kondopoulou and Westphal (1986)
6. Axios (Oreokastro)	Plutonics	Eocene	1	29.5	39	–	Feinberg et al. (1994)
7. Axios (Goumennisa)	Plutonics	Eocene	1	89	49	14	Feinberg et al. (1994)
8. Chalkidiki (Paliouri)	Sediments	Jurassic	9	32.1	50	18	Feinberg et al. (1994)
9. Chalkidiki	Ophiolites	Jurassic	5	314	34	13	Edel et al. (1991)
10. Axios – Chalkidiki	Ophiolites	Jurassic (?)	16	318.7	36.5	9.5	Feinberg et al. (1996)
11. Chalkidiki (Monopigado)	Granites	Jurassic	1	56.5	25	16	Feinberg et al. (1996)
Expected declination and inclination values for the area							
Kassandra		Eocene		10.4	50.8		Westphal et al. (1986)
Kassandra		Eocene		10.1	53.9		Besse and Courtillot (2002)
Kassandra		Oligocene		8.2	56.4		Westphal et al. (1986)
Kassandra		Oligocene		10.3	56.1		Besse and Courtillot (2002)

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Table 3. (b) Published data from the Mesohellenic basin.

Area	Age	N	D	I	k	A_{95}	Reference
Mesohellenic Basin EE (mean)	Oligocene-M.Miocene	6	51.7	28.2	19.5	15.5	Van Hinsbergen et al. (2005)
Mesohellenic Basin EE (mean)	Eocene – Oligocene	4	65.4	39.1	15.3	24.3	Van Hinsbergen et al. (2005)
Mesohellenic Basin DD (mean)	Oligocene – M.Miocene	8	51.0	20.3	25.8	11.1	Van Hinsbergen et al. (2005)
Mesohellenic Basin Krania (mean)	L. Eocene	6	46.4	19.5	27.2	13.1	Van Hinsbergen et al. (2005)
Mesohellenic Basin SM (mean)	36–24 Ma	5	27.0	47.0	16.0	10.0	Kissel and Laj (1988)

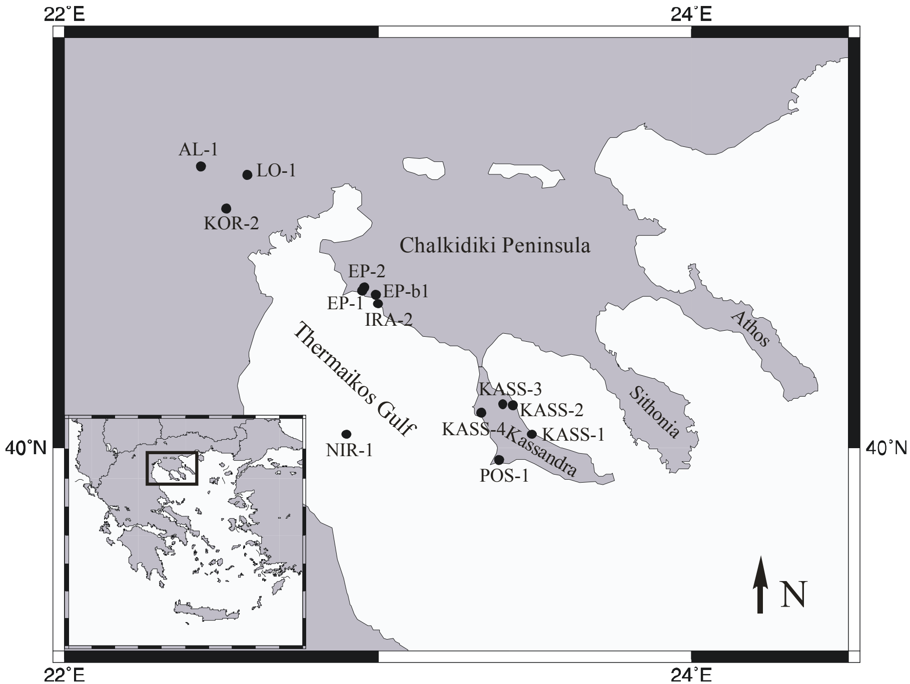

Fig. 1. Schematic map of Chalkidiki Peninsula, where the location of the studied drill cores is shown.

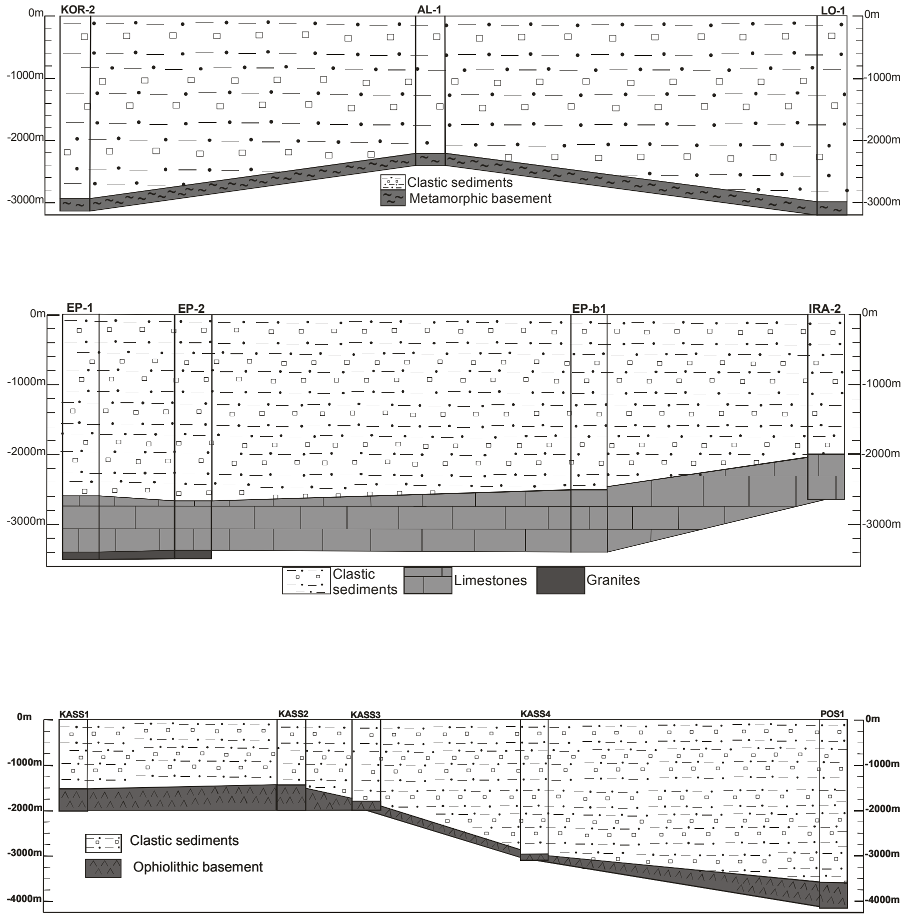

Fig. 2. Lithostratigraply of the studied cores.

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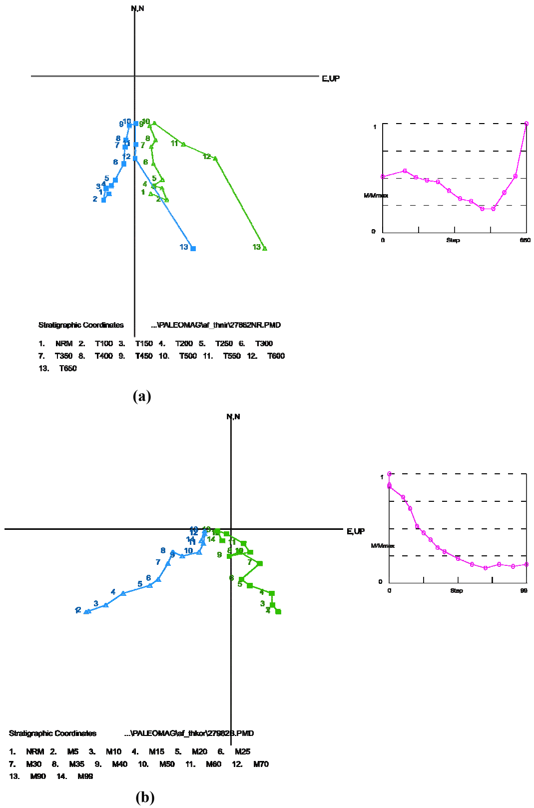

Fig. 3. Zijderveld diagrams and normalized magnetic intensity curves with stepwise AF thermal **(a)** and AF **(b)** demagnetization for representative samples. Open squares: projection onto the vertical plan, Black squares: projection onto the horizontal plan T100: Thermal demagnetization at 100°C, M10: AF demagnetization at 10 mT.

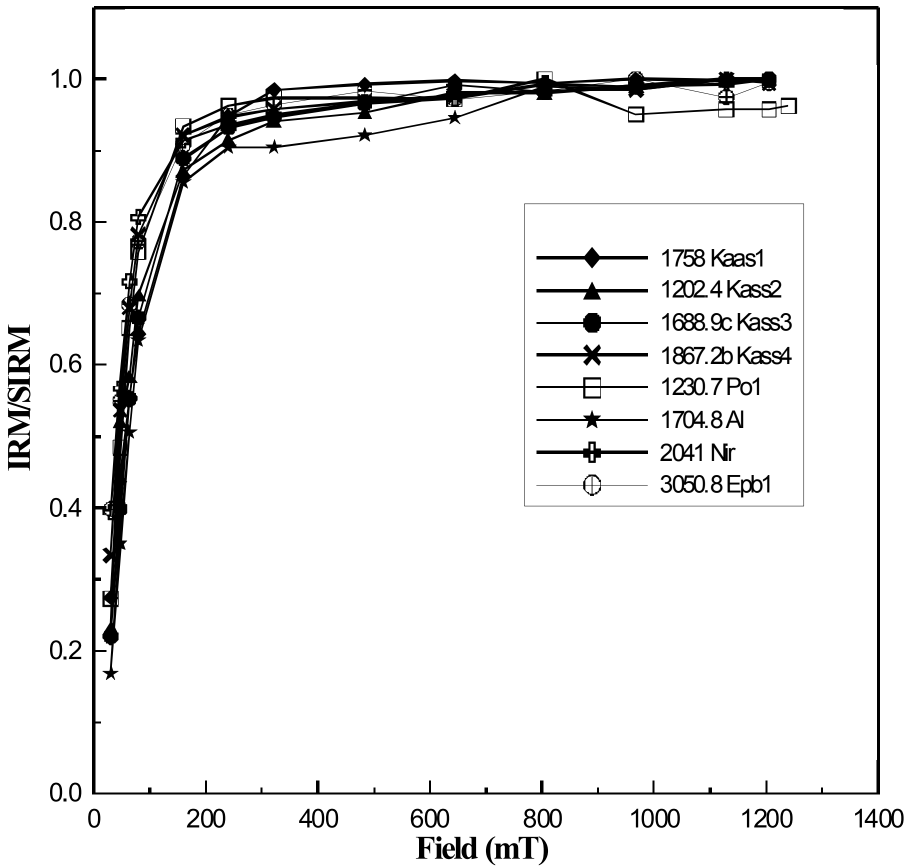

Fig. 4. IRM acquisition curves for representative samples.

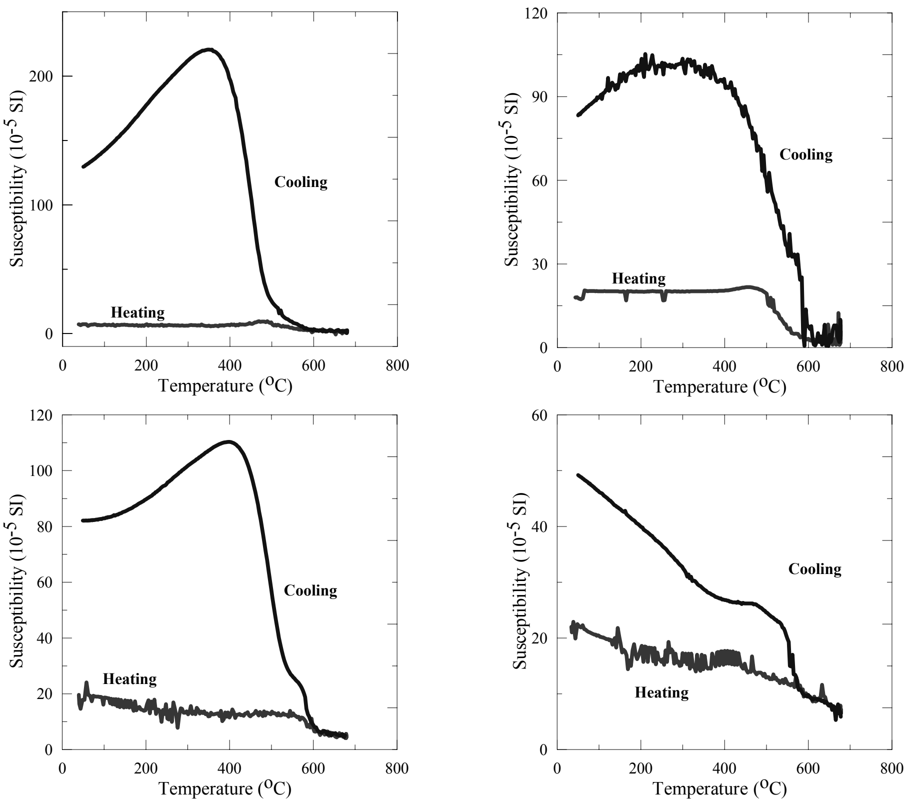

Fig. 5. Thermomagnetic analysis indicating the presence of magnetite and the transformation of low susceptibility to high susceptibility minerals during cooling.

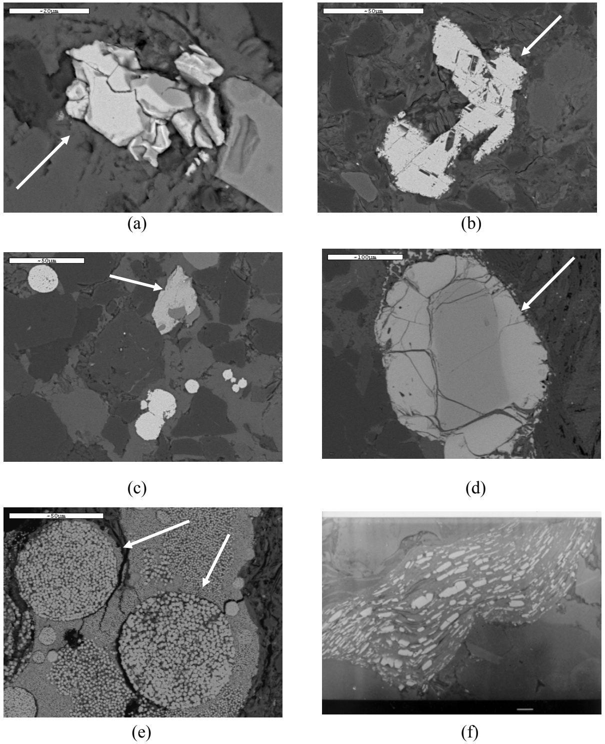

Fig. 6. **(a)** Scanning electron micrograph of a magnetite crystal, **(b)** Scanning electron micrograph of an hematite crystal **(c)** Scanning electron micrograph of an ilmenite crystal, **(d)** Scanning electron micrograph of chromite crystal, **(e)** Backscattered electron micrograph showing framboids in colonies, surrounded by a thin surface membrane, **(f)** Scanning electron micrograph showing pyrite microcrysts of different size, not organized in framboids.

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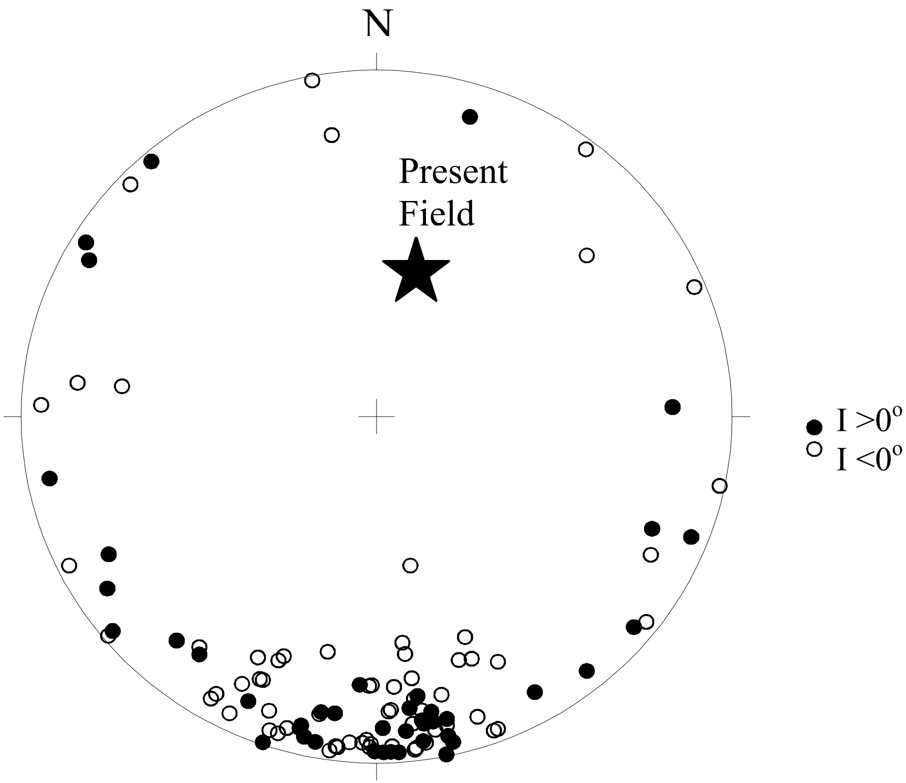

Fig. 7. Stereographic projection of the isolated viscous component of the samples showing very shallow inclination values.

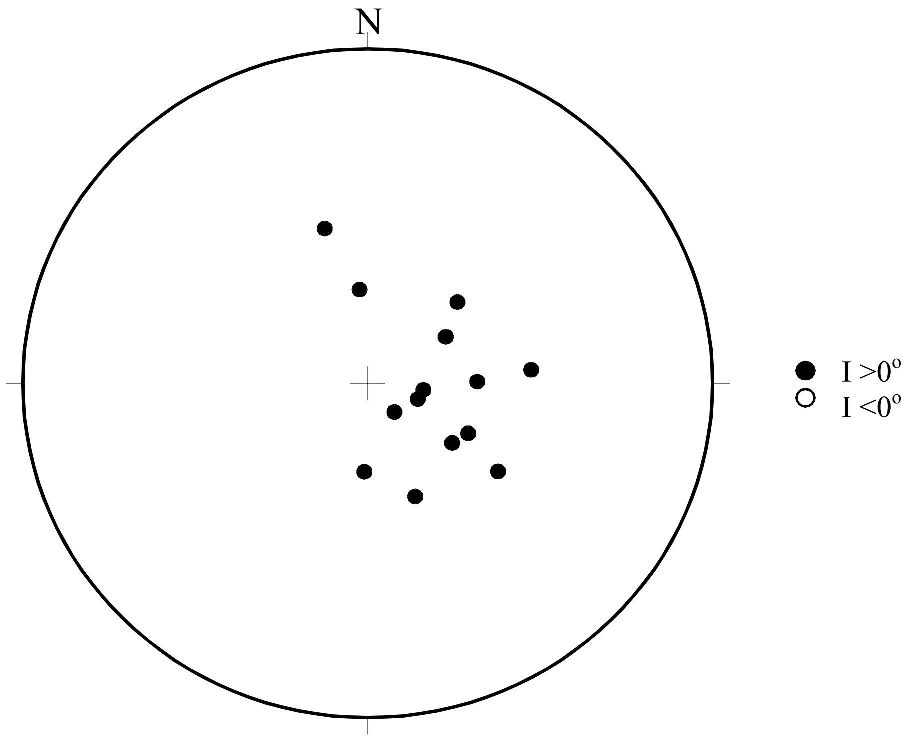

Fig. 8. Viscous component of Axios samples.

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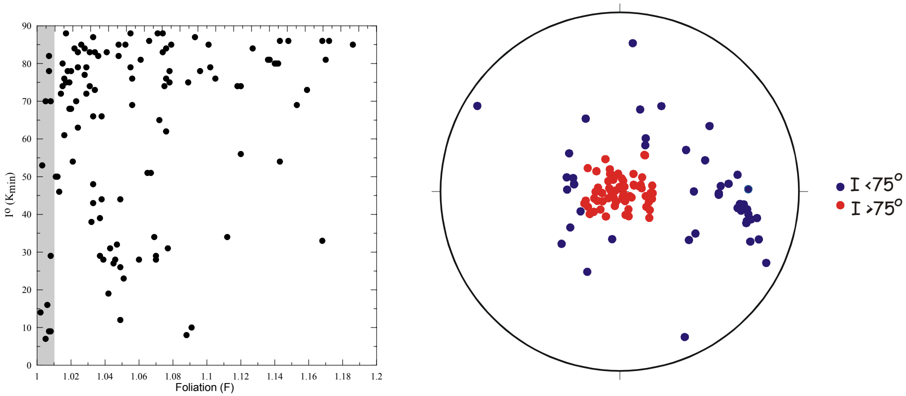

Fig. 9. Inclination values versus foliation (left) and stereographic projections of the direction of K_{min} axe of anisotropy of the samples (right).

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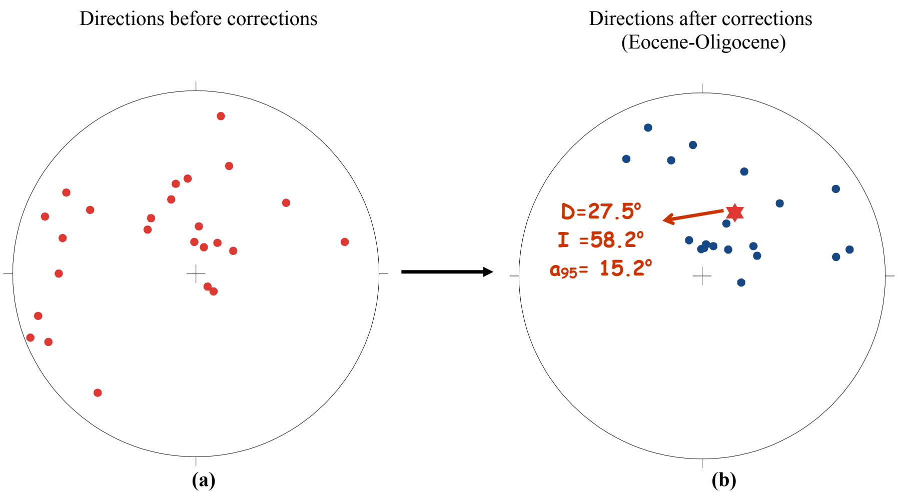

Fig. 10. **(a)** Initial directions of the samples before the reorientation, **(b)** final directions after the corrections. Red star represents the mean direction of these samples.

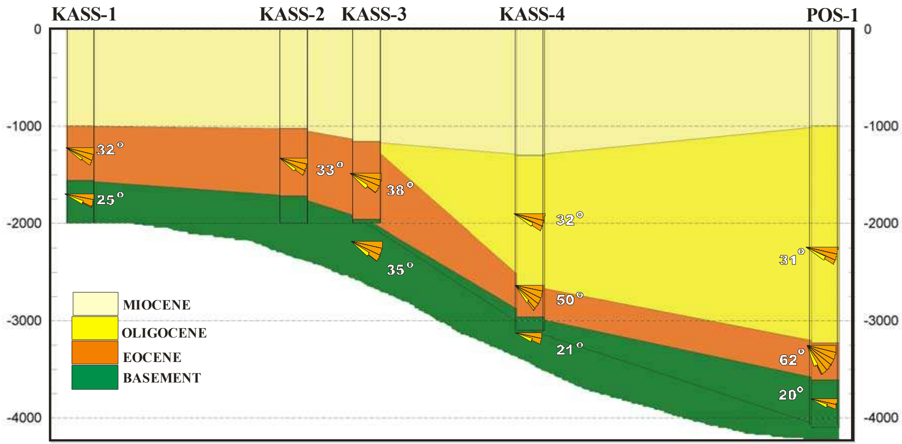

Fig. 11. Distribution of inclination values for Kassandra area. Orange funs represent the inclination values, since the yellow ones the a_{95}.